

Hydrodynamic modelling of tidal-fluvial flows in a large river estuary

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ABSTRACT

The transition between riverine and estuarine environments is characterised by a change from unidirectional to bidirectional flows, in a region referred to herein as the Tidally-Influenced Fluvial Zone (TIFZ). In order to improve our understanding of the hydrodynamics and morphodynamics of this zone, we present a combined field and numerical modelling study of the Columbia River Estuary (CRE), USA, tidally-influenced fluvial zone. The CRE is large measuring 40 km in length and between 5 and 10 km wide. A shallow water model (Delft3D) was applied in both 2D and 3D configurations and model sensitivity to the key process parameterizations was investigated. Our results indicate that a 2D model constrained within the estuary can sufficiently reproduce depth-averaged flow within the TIFZ of a stratified estuary.

Model results highlight the interactions between tidal-, fluvial- and topographic-forcing that result in depth dependent tidal rectification, and thus zones of residual sediment transport that: i) may be flood-directed along shallow channel margins and in the lee of bars, and simultaneously ii) is ebb-directed within deeper channel thalwegs. This condition is enhanced at lower discharges, but increased fluvial discharge reduces the number and size of regions with net flood-directed sediment transport and flow. These sediment transport patterns provide a mechanism to extend the bar/island topography downstream, and generate flood-directed, ebb-directed, and symmetrical bedforms, all within the same channel. Analysis of the model data reveals flood-directed sediment transport is due to both tidal variability and mean flow. These results highlight the need to include the mean flow component (M_0) when considering the long-term morphodynamic evolution in a TIFZ.

1. Introduction

Tidally-influenced fluvial zones (TIFZs) represent some of the most complex environments on Earth (Dalrymple and Choi, 2007), and extend from the mouth of an estuary, or delta, through the freshwater tidal river reach, whose landward limit terminates at the most upstream point of observable variations in water surface elevations caused by tidal-flows (Hoitink and Jay 2016). TIFZs are commonly sub-divided into three general hydraulic zones defined by the mean ratio of tidal vs fluvial energy input, and include (see Jay et al., 1990; Jablonski et al., 2016): i) tidally-dominated lower river (i.e., estuary), ii) tidally-dominated, fluvially-influenced, or mixed tidal-fluvial regime, and iii) fluvially-dominated, tidally-influenced regime. Thus, it is at the boundary between the tidally-dominated lower TIFZ, or brackish water upper

estuary, and the mixed tidal-fluvial regime (downstream boundary of freshwater tidal river reach; middle TIFZ), where tidal-flows are strongly impacted by varying river discharges, bed geometry/bathymetry, and water density (Jay and Smith, 1988). These factors drive asymmetric hydrodynamic flows with either flood- or ebb- dominance (Dyer 1997, chapter 2), and associated net sediment transport (Guo et al., 2014), and are referred to as tidal rectification (cf. Huthnance 1973; Bowers and Al-Barakati 1997).

The hydrodynamics and morphodynamics in tidally-dominated estuaries have been well studied (cf. Guo et al., 2014; Zhou et al., 2014; Chu et al., 2015; Guo et al., 2016), with net sediment transport related to: i) tidal velocity asymmetries (Van Maren et al., 2004; Díez-Minguito et al., 2012; Guo et al., 2014); ii) the generation of the M_4 overtide (Friedrichs and Aubrey, 1988); and iii) the triad interaction M_2 - O_1 - K_1

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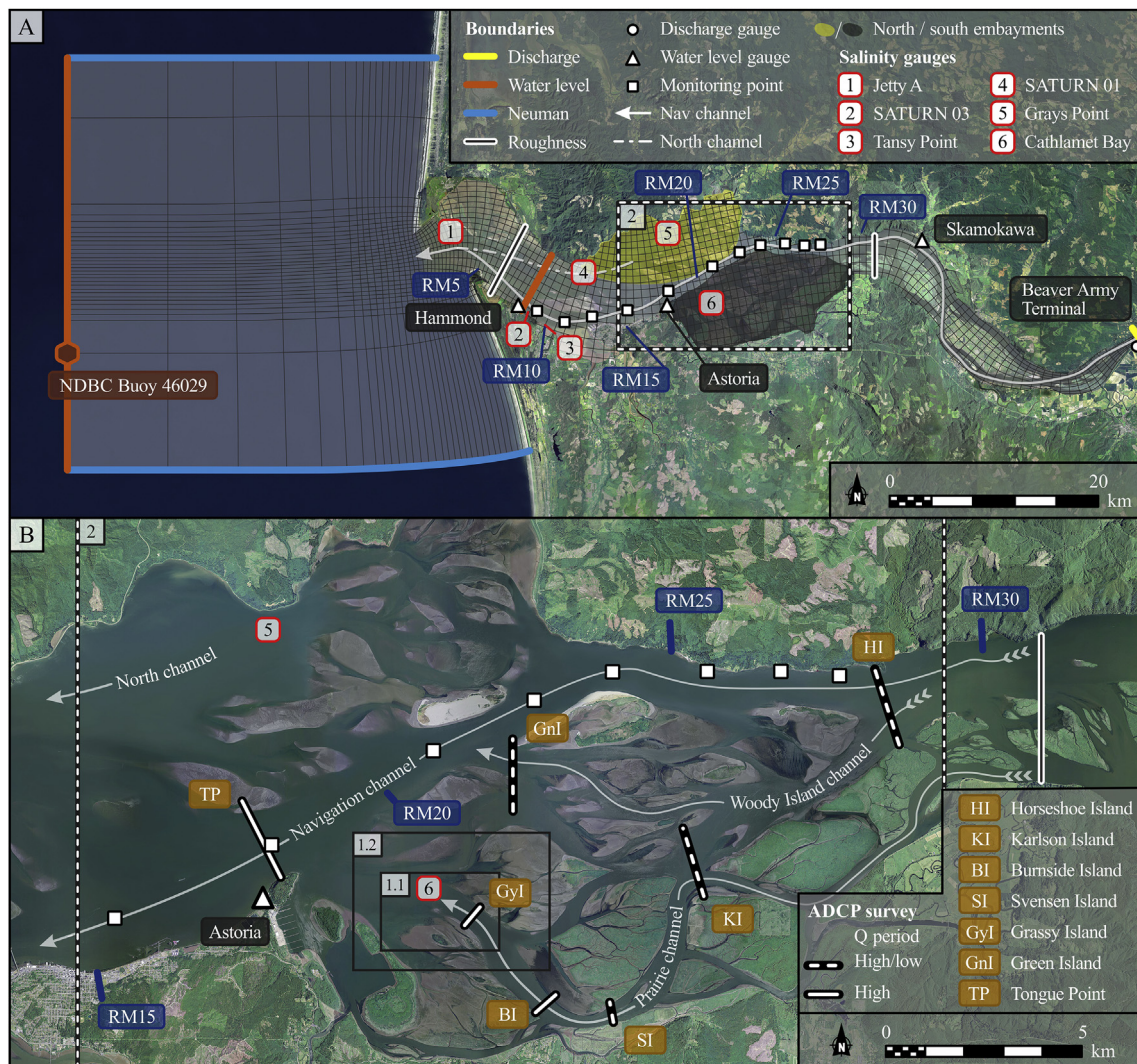


Fig. 1. Aerial photography of the lower Columbia River Estuary showing (A) modelling domain location and boundaries and, (B) ADCP cross-section locations. Model grid resolution has been reduced by a factor of 8 for clarity. Aerial photographs (2009) provided courtesy of the United States Department of Agriculture National Agriculture Imagery Program.

(Hoitink et al., 2003). The focus of recent work on the middle to upper TIFZ includes suspended sediment dynamics (Achete et al., 2016), interactions between mean and tidal flow components (Van Maren et al., 2004; Chu et al., 2015), net water transport (Alebregeetse and de Swart, 2016), and determination of the relative importance of non-stationary river vs stationary tidal energy signals in governing streamwise fluctuations in the water surface elevation (Matte et al., 2013; Guo et al., 2015; Jay et al., 2015), and/or flow velocity magnitudes and orientations (Losada et al., 2017).

A number of process-based hydraulic models have been applied to study the Columbia River Estuary (CRE) (Kärnä and Baptista, 2016), including the Regional Ocean Modelling System (Liu et al., 2009), EL-CIRC (Zhang et al., 2004), SELFE (Zhang and Baptista, 2008) and Delft3D (Elias et al., 2012). The predominant focus of these model applications has been on plume dynamics and estuary mouth processes. Because of this focus, the impact of stratification was important and thus these models were all deployed in 3D mode.

Modelling studies of other estuaries include 3D (e.g. Burla et al., 2010) and 2D (e.g. Bolle et al., 2010; Prario et al., 2011) representations. In most cases, the modelling domains include the estuary mouth and so apply a coastal boundary condition. Notable exceptions are Prario et al., (2011) and Matte et al., (2017a, 2017b), who constrained their models to within the estuary. This suggests that it may be possible

to apply a 2D model to the TIFZ of the CRE. Herein, we test this by applying both 2D and 3D simulations with the model boundary located both inside and outside the estuary.

Despite this recent interest, little is known about TIFZ channel scale hydrodynamics and morphodynamic processes that determine how channels and sand bars evolve, which ultimately are one of the important sedimentological building blocks of estuaries and deltas (Dalrymple and Choi, 2007; Leuven et al., 2016). Therefore, to investigate TIFZ channel to barform evolution, a combined field and numerical modelling study was conducted on the meso-tidal lower Columbia River (LCR), WA/OR, USA (Prokocki et al., 2015).

Thus, in this paper three principal questions are addressed:

1. Can a 2D model be used to simulate accurately the LCR TIFZ (i.e., estuary) where density stratification is present?
2. What are the dominant characteristics of LCR TIFZ channel scale hydrodynamics and morphodynamics, and how do these processes affect bedform to barform evolution?
3. How do fluctuations between low- and high- river flow impact LCR TIFZ bedform to barform evolution?

To address these points, a modelling study of the LCR TIFZ including different levels of process representation to evaluate the

impacts of waves, 3D/2D, boundary location, and bedform roughness and turbulent viscosity, is developed herein. Details of the modelling setup are provided in section 3. The model that performed best was assessed by calibration, using measured water level data, and validation with other hydrodynamic data (sections 4.1 and 4.2). Results from the best performing model were then used to investigate channel-scale hydrodynamics and morphodynamics within the LCR TIFZ (section 4.3). Due to the large variability in river-discharge, the effect of flow discharge on model calibration/validation and channel scale process was also analysed.

2. Columbia River estuary

The Columbia River is the largest river entering the Pacific Ocean from North America, with an average river-discharge of $\sim 7300 \text{ m}^3 \text{ s}^{-1}$ (Jay et al., 2010). The discharge is seasonally variable with a typical minimum autumn-early spring flow of $\sim 3000 \text{ m}^3 \text{ s}^{-1}$ and a spring freshet between 10,000 and 15,000 $\text{m}^3 \text{ s}^{-1}$. The tide is mixed diurnal and semidiurnal with a tidal range of 3.6–4 m (Fain et al., 2001). Within the Columbia River littoral cell, the wave climate is seasonally-variable and high-energy with significant wave heights $H_s = 3 \text{ m}$ and peak period $T_p = 12 \text{ s}$ in the winter season (Gelfenbaum and Kaminsky, 2010). A deep thalweg (navigation channel), which is periodically dredged, meanders through the central portion of the estuary flanked to the north and south by embayments (Fig. 1).

2.1. Data availability

Bathymetric and topographic measurements from various sources and time periods are available offshore and within the estuary. The National Oceanic and Atmospheric Administration (NOAA) collected Multi-Beam Echo Sounding (MBES) data predominantly within the Navigation Channel stretching from the mouth to river-mile 18 (2007), river-mile 18 to 25 (2008), river-mile 25 to 28.5 (2011) and from river-mile 28.5 to Beaver Army Terminal (2009). Areas outside the Navigation Channel were surveyed using Single-Beam Echo Sounding (SBES) by the United States Army Corps of Engineers (USACE) in 2003 and 2009. Shallow areas inaccessible to the USACE vessels were surveyed using SBES in 2009 and 2010 by the Lower Columbia River Estuary Partnership (LCREP). Exposed topography within the estuary was measured by Light Detection and Ranging (LiDAR) in 2005 by the Puget Sound LiDAR Consortium (PSLC). Just outside the estuary mouth, NOAA (2007) and USACE (2009) conducted MBES bathymetric surveys. For the remainder of the coastal zone, 20 km north and south and 40 km west of the mouth, data collected by NOAA using a fathometer during the 1920s and 1950s are also available.

Gauging stations maintained by NOAA and CMOP (Coastal Margin Observation & Prediction) are present throughout the system, collecting both hydrological and biological data. Hydrological data monitored in the estuary include water surface elevation, salinity, temperature, wave climate and discharge. The data used in this paper were obtained from CMOP (<http://www.stccmop.org>), NOAA (<http://www.noaa.gov>) and USGS (<http://waterdata.usgs.gov>). The locations of the gauging stations are shown in Fig. 1A.

2.2. Flow surveys

Flow data were obtained using a Teledyne RD Instruments 1200 kHz ADCP, deployed from a small inflatable boat, which was set to determine three-dimensional flow in a vertical column at 0.25 m interval bins at a rate of $\sim 1 \text{ Hz}$. The ADCP was deployed in moving-boat mode with vessel position and velocity correction provided via Real Time Kinematic GPS. The distance from the water surface to the first ADCP measurement bin was 0.91 m, which accounted for submergence of the ADCP below the water-line and the blanking distance that removes the effect of acoustic ringing (Szupiany et al., 2007). Data from the bottom

6% of the profile were also removed in processing, to remove the effect of contamination by side-lobe interference. Due to the large length scales and short time-scales for survey data collection, it was not possible to take repeat measurements for stable flow conditions. Therefore, in this paper, velocity data are depth-averaged to remove non-stationary turbulent signals.

The average boat velocity during the surveys was 1.09 m s^{-1} , providing high-resolution (~ 6.3 measurements m^{-2}) velocity data along each cross-section. Seven transects were surveyed during the high river-discharge period (field seasons 1 and 3). A subset of these transects were re-surveyed during the low river-discharge period (field-season 2). The locations of all the surveyed transects are illustrated in Fig. 1. The transects surveyed at low river flow are marked as dashed lines. The data were processed using the VMT Matlab toolbox (Parsons et al., 2013). Raw and processed data files are available at: <http://fluvialtidal.com>.

The field survey collection periods ranged from 4 to 8 h per deployment with an average of 5 h depending on the prevailing conditions. Each transect was typically surveyed at least five times. The typical time for each survey was around 30 min but smaller channels could be crossed in 10 min while others took approximately an hour to survey. Processing of ADCP data using a time-interpolation method produced no perceptible difference in the spatial distribution of velocity.

3. Modelling simulations

3.1. Numerical model

Both 2D and 3D Delft3D-FLOW (Deltares, 2011) models were setup to simulate the Columbia River TIFZ. The horizontal momentum equations include forcing associated with: i) Coriolis force (model latitude $\phi = 46.1881^\circ$); ii) horizontal and vertical kinematic eddy viscosity ν_v and ν_H ; iii) water-surface and bed shear stresses; and iv) pressure gradient forces expressed through the application of a hydrostatic pressure assumption.

Free surface and bedform shear stresses are included using the quadratic friction law: $\tau_b = \rho_{ref} C_d \vec{U}_{ref} |\vec{U}_{ref}|$, where τ_b is the shear stress, ρ_{ref} is the reference density, C_d is a drag coefficient and \vec{U}_{ref} is the reference velocity. At the free-surface, the air density and wind velocity are used in this formulation and C_d varying between 0.0025 and 0.0289 (Elias et al., 2012). For bed roughness effects, a slightly modified version of drag coefficient is used; the Chezy roughness parameter: $C = \sqrt{g/C_d}$. Also, in this case the water density and the near-bed (3D model) or depth-averaged (2D model) velocities are used (\vec{U}_{ref}) in the shear stress parameterisation.

The viscosity associated with vertical shear is: $\nu_v = \nu_{mol} + \max(\nu_{3D}, \nu_v^{back})$, where ν_{mol} is the molecular fluid viscosity, ν_{3D} is the 3D eddy-viscosity and ν_v^{back} is the vertical background eddy-viscosity. In the horizontal, the viscosity is: $\nu_H = \nu_v + \nu_H^{back} + \nu_{SGS}$; where ν_H^{back} is the user specified horizontal background eddy-viscosity and ν_{SGS} is the horizontal Sub-Grid-Scale (SGS) eddy-viscosity. The SGS terms are computed using the Uittenbogaard HLES (Horizontal Large Eddy Simulation) model (Deltares, 2011, pp. 532–537). A similar relationship holds for the diffusive terms (D_v and D_H) (Deltares, 2011, pp. 203). In the present paper, the background viscosity and diffusion are defined following Elias et al., (2012) whereby $\nu_H^{back} = D_H^{back} = 1 \text{ m}^2 \text{ s}^{-1}$ and $\nu_v^{back} = D_v^{back} = 10^{-6} \text{ m}^2 \text{ s}^{-1}$. The 3D viscosity and diffusion (ν_{3D} and D_{3D}) are computed using the $k-\epsilon$ turbulence model (Deltares, 2011, pp. 231–233). The cyclic advection scheme is used for both momentum and scalar transport.

In some of the simulations (Table 1), the hydrodynamic model was coupled with Delft3D-WAVE (SWAN version 40.72ABCDE). The model parameterisation of Elias et al., (2012) is used herein, including an overlapping WAVE grid.

Table 1

Modelled time periods and configurations assessed including a summary of the optimal roughness parameters.

Modelling periods and mean fluvial discharge					
Field season	Mean fluvial discharge (m ³ s ⁻¹)			Model period	
1	15,700			27/05/2011–27/06/2011	
2	4200			15/09/2011–15/10/2011	
3	11,500			29/05/2012–29/06/2012	
Configurations					
Case	2D/3D	fx2 (m2s-1)	Waves	Csea (m1/2 s-1)	Criv (m1/2 s-1)
1	3D	1.0	–	70	60
2	3D	1.0	Included	70	60
3	2D	1.0	–	60	60
4	2D	1.0	Included	60	60
5	2D	0.1	–	60	60
6	2D	10.0	–	60	65*
7	2D	HLES	–	60	65

3.2. Model application

3.2.1. Domains

Two domains were created (see Fig. 1A). The first covered an area similar to the modelling domain used by Elias et al., (2012), and included the Columbia River (~ 30 km long and ~ 3 km wide), Estuary (~ 40 km long and ~ 5 – 10 km wide) and coastal zone (~ 40 km perpendicular and parallel to the shore). The second domain was a reduced version of the first, where the downstream boundary was moved from outside to inside the estuary (Fig. 1A).

A variable resolution grid was used ranging from ~ 80 m in the estuary and up to 1000 m in the coastal region. In the 3D model simulations, the resolution was reduced by a factor of two to facilitate acceptable model run-times. The vertical resolution was greatest near the bed and free surface (4% of the depth) and lowest in the middle (17% of the depth), varying smoothly in-between over 10 layers. For the remainder of this paper, these domains are referred to as the sea-boundary and estuary-boundary domains respectively.

2D and 3D models were run in the estuary- and sea-boundary domains respectively. We initially tested a 2D model in the sea-boundary domain but the performance was poor. We chose not to implement a 3D model in the estuary-boundary domain because it was not possible to obtain accurate salinity boundary conditions.

3.2.2. Boundary conditions

The models were forced using discharge and water level boundary conditions. Water level data from Hammond tide gauge was used for the estuary-boundary simulations. An astronomical water level was specified at the western boundary of the domain in the sea-boundary simulations. The amplitude and phase of the nine components used were obtained from Elias et al., (2012).

The Local Mean Sea Level (LMSL) was specified using a conversion from NAVD88 to LMSL provided by NOAA's VDatum software (vdatum.noaa.gov). The water level boundary was supplemented with two zero-normal water level gradients (Neumann) at the north and south boundaries (Fig. 1A). Time-series discharge observations obtained from the Beaver Army Terminal gauging station were used for the river boundary. This data includes the effects of both river- and tidal- forcing. The wind climate was specified using measured data from NOAA (NDBC buoy 46,029; see Fig. 1A for location) and the same parameterisation as Elias et al., (2012).

The salinity at the river and the sea boundaries during inflow were set equal to 0 and 33, respectively (Elias et al., 2012). A Thatcher-Harleman time-lag (Deltares, 2011) of 120 min was applied to provide smoothing.

3.2.3. Parameters

A series of model simulations were conducted to calibrate and assess model sensitivity. This assessment focused on: bed roughness; waves; model dimensionality; horizontal viscosity; and downstream boundary location.

The roughness values were specified spatially variable with: uniform roughness in the lower estuary and coastal domain (C_{sea}); uniform roughness in the river (C_{riv}); and a linear variation in the intervening region (see Fig. 1A for transition points). This follows similar approaches by others (Baptista et al., 2005; Sassi et al., 2011; Elias et al., 2012; Van et al., 2016). The range of values assessed was based on the values used in other studies of the CRE (Baptista et al., 2005; Elias et al., 2012).

A series of simulations were conducted using the baseline parameters described in section 3.2 for FLOW and FLOW-WAVE configurations. A further three model configurations were considered for the 2D estuary-boundary simulations where the horizontal background viscosity is varied (0.1 , $10 \text{ m}^2 \text{s}^{-1}$ and HLES). For each of these configurations, one-month long simulations were run (Table 1).

4. Results

4.1. Model calibration and sensitivity

The Mean Tidal Amplitude difference $\text{MTA}_d = (\text{MTA}_m - \text{MTA}_o)$ and Mean Water Level difference $\text{MWL}_d = (\text{MWL}_m - \text{MWL}_o)$ are used to calibrate the models. The subscript 'm' and 'o' indicate model predictions and observations. Figure 2 shows filled contour plots of MWL_d averaged across all available gauges for both field seasons two and three. The dotted line corresponds to the zero-contour for MTA_d ($=\text{MTA}_{d0}$). The optimal configuration is obtained where the zero-contour for both MWL_d ($=\text{MWL}_{d0}$) and MTA_d intersect. The 3D model over-predicts the MWL and so there is no zero-contour (Fig. 2A and B). The results indicate that the model is more sensitive to tidal amplitude. For both sets of the 3D simulations, $C_{\text{riv}} = 60 \text{ m}^{1/2} \text{s}^{-1}$ and $C_{\text{sea}} = 70 \text{ m}^{1/2} \text{s}^{-1}$ provide optimal results (values are larger than Elias et al., 2012). The optimal roughness values are summarised in Table 1.

The 3D sea-boundary model simulations (Fig. 2A and B) show that the wave model increases the mean water level by around 0.025 m but has a negligible effect on tidal amplitude. In the 2D estuary-boundary simulations (Fig. 2C and 2D) the wave model has a negligible effect but viscosity can lead to a significant shift in the tidal water levels (Fig. 2C and F).

An inversion of the roughness sensitivity of the MWL metric to the sea roughness parameter is evident for the 3D model but not the 2D model (not shown). The mean high and low water levels (Fig. 3A–D) show that the model is sensitive to the applied C_{sea} at high water during low river-discharge, but insensitive during high-river-discharge (Fig. 3A and B). The reverse is true for the mean low water level (Fig. 3C and D).

The response of the model at high and low water levels is related to flood and ebb processes, and in particular to the work done by external forces. Considering only the bed shear forcing ($F \propto U^2$), the work done $W (=FUDt)$ may be computed and partitioned into ebb (e) and flood (f) periods:

$$w_d = (W_e - W_f)/(W_e + W_f), \quad (1)$$

There is an increase in w_d in the upstream direction, as well as a clear difference in the work done during the periods of high and low river-discharge. During low river flow, the work done in the flood period dominates in the lower to middle part of the estuary, corresponding to the increased sensitivity to roughness at high water levels (flooding process). During high river flow, the ebb work-done dominates for most of the estuary, corresponding to the increased sensitivity to roughness at low water levels (ebb process).

A direct comparison between 2D estuary-boundary and 3D sea-boundary results (Fig. 4) is presented relative to the Hammond water

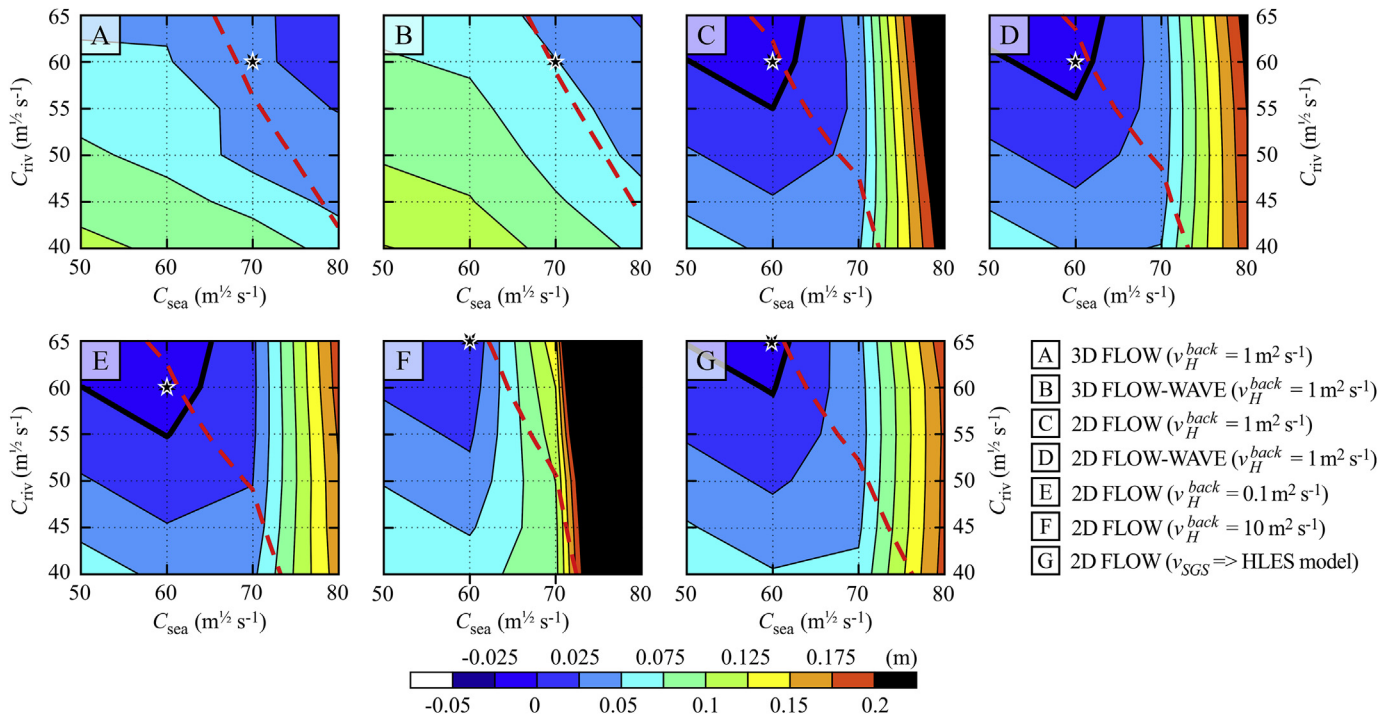


Fig. 2. Filled contour plots of MWL_d averaged across all available gauges and river flow periods (field-seasons 2 and 3). Red dotted line is the zero-contour for MTA_d . (For interpretation of the references to colour in this figure legend, the reader is referred to the Web version of this article.)

level data. These results demonstrate that the 2D estuary-boundary model performs best and this is most noticeable at the Skamokawa tide gauge.

4.2. Model validation

The salinity data were assessed using the Root Mean Square Error (RMSE), and the model skill using the Index of Agreement (Willmott, 1981). The results from this analysis are not presented here but they demonstrate: i) neither model configuration (FLOW or FLOW-WAVE) systematically out-performed the other; ii) the model skill is higher in the lower part of the estuary, a result also observed by Elias et al., (2012); and iii) the present model performs to a similar level of accuracy as other modelling studies of the CRE (Elias et al., 2012; Kärnä et al., 2015; Kärnä and Baptista, 2016).

Although it is possible to obtain the non-stationary amplitude and phase (Matte et al., 2013), herein we use the t_{tide} Matlab toolbox (Pawlowicz et al., 2002) since the analysis window is one month. The 2D and 3D configurations (case 1 and 3: Table 1) and observations were analysed (Table 2).

Both model configurations reproduce the tidal amplitudes to a similar level of accuracy: the average and maximum amplitude errors being 0.01 m and 0.05 m respectively. The 2D estuary-boundary model performs slightly better at reproducing the tidal phase with an average/maximum phase error of around $3/5^\circ$ compared to $7/12^\circ$ for the 3D sea-boundary configuration. The results also show that the amplitudes are damped by increasing river discharge for M_2 , S_2 and O_1 but amplified for K_1 . The components that are affected most are K_1 and S_2 .

We have also computed the model skill and absolute error (Table 3). The results show that the RMSE is relatively large for the 3D case. This is surprising given the amplitude and phase differences between these configurations is small. This is due to poor representation of the MWL as discussed in section 4.1 and demonstrated in Fig. 2A and B.

In Fig. 5A–D, a direct comparison between the modelled and measured velocity magnitude is given for four ADCP transects with error bars for variability. The degree of bias is illustrated using the line fitted

using Reduced Major Axis (RMA) linear regression (compare with the 1:1 line of equality). In Fig. 5E, the RMA gradient for each transect and model configuration is shown. The average of the absolute difference between 1 and the computed gradient ($|1 - \bar{m}|$) across all transects is shown in Fig. 5F. The degree of scatter is quantified here using the correlation coefficient r (Fig. 5G). The mean differences between a perfect correlation coefficient of 1 and the actual correlation coefficients ($1 - \bar{r}$) are presented in Fig. 5H. These results show that the 2D estuary-boundary model with uniform viscosity $\nu_H^{\text{back}} = 1 \text{ m}^2 \text{ s}^{-1}$ performs best.

4.3. Hydrodynamics and morphodynamics in the TIFZ

The 2D estuary-boundary model results (from configuration 3 in Table 1) were used to investigate the channel scale hydrodynamics and morphodynamics in the TIFZ of the CRE. This is the best performing configuration based on our analysis. To begin, small regions (1.1 and 1.2, Fig. 1) are used to investigate temporal flow patterns, total sediment transport and the contribution of tidal and fluvial forcing. These results are then used to generalise the results in the upper part of the CRE (region 2, Fig. 1).

4.3.1. Temporal flow patterns in region 1.1

The flow patterns during a high discharge period (field season 3, Table 1) where the flow changes from ebb to flood in region 1.1 (Fig. 1B) are presented in Fig. 6. This region includes the Grassy Island ADCP transect (Fig. 1B). Measured and modelled data from three time periods are shown in Fig. 6G–I.

In Fig. 6A, the velocities are ebb-directed and approximately at maximum magnitude. During the next 2 h, the velocity magnitude is reduced (Fig. 6B), and 1 h later (Fig. 6C) the flow remains predominantly ebb-directed, although some of the flow has reversed and a recirculation zone develops.

In the following hour, nearly all the flow has reversed (Fig. 6D). In the Prairie Channel (Fig. 1B), the flow is flood-directed in the shallow areas but remains ebb-directed in the deeper part of the channel. This

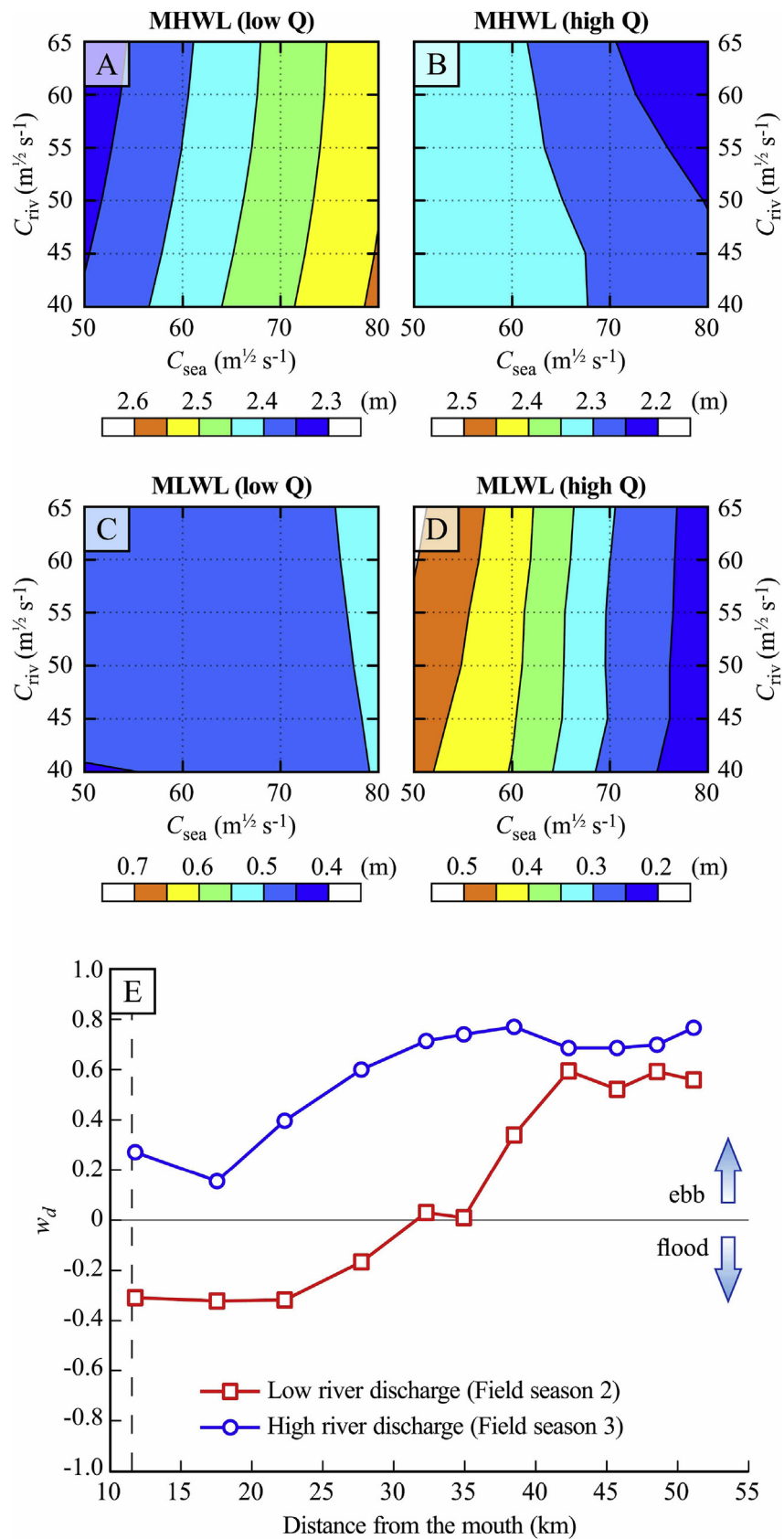


Fig. 3. (A–D) Mean high and low water levels predicted by the 3D FLOW-WAVE model. (E) Proportion of the total work done during flood and ebb calculated using Equation (1). Vertical dash-line denotes Hammond tide gauge. Positive w_d indicates ebb is dominant, and negative w_d indicates flood is dominant (Fig. 3E).

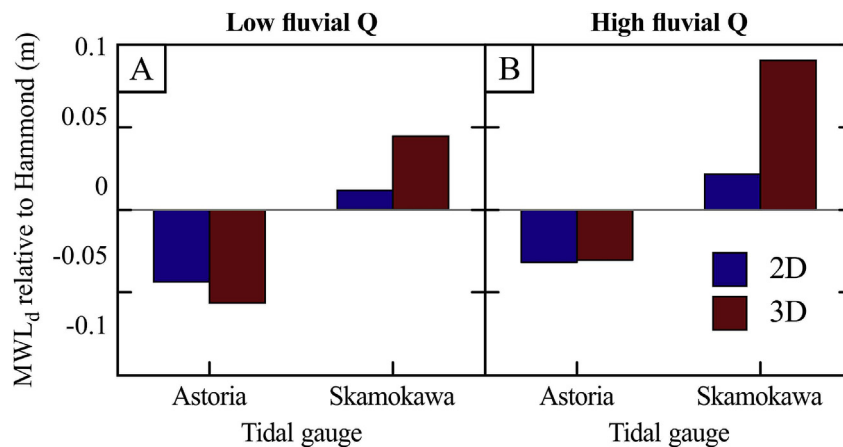


Fig. 4. Difference between predicted and measured mean water level.

bidirectional flow pattern is exhibited in both the measured and modelled data (Fig. 6G) and the recirculation zone is still evident (Fig. 6D). The recirculation is driven by opposing ebb-directed flow from the Prairie Channel and flood-directed flow from downstream. Thirty minutes later (Fig. 6E), the flow in this area is entirely ebb-directed, with the velocity magnitude in the shallow areas being much greater than in the deeper part of the channel. The flow profiles illustrate this also (Fig. 6H). Finally, 1 h later (Fig. 6F and I), the flow velocities in the deeper part of the channel are now greater than in the shallow areas.

The flow in this region is affected by both tidal and fluvial flows, as well as a clear influence of bed topography. Analysis of the results showed that the flow duration is flood-biased in some areas of this region, typically along the margins of the channel and in the lee of Grassy Island (with respect to the ebb direction). The same general flow patterns are observed during the low discharge period although the magnitudes of the ebb flow velocities and duration are reduced.

4.3.2. Sediment transport in region 1.2

The influence of tidal and fluvial forcing, combined with bathymetric variability, may lead to mean flood-directed flow. This is demonstrated for a slightly larger area (region 1.2, Fig. 1B) than outlined in the previous section. To compute the mean flow direction, the model was run in a steady configuration using mean water level height and discharge at the model boundaries. This was used to define a directional parameter d ($= \pm 1$) to denote ebb- ($d = +1$) or flood- ($d = -1$)

Table 3

Index of agreement and root mean square error computed for the low (field season 2) and high (field season 3) discharge periods.

Gauge	IOA		RMSE (m)					
			field-season 2		field-season 3			
	2D	3D	2D	3D	2D	3D	2D	3D
Ham	–	0.979	–	0.979	–	0.22	–	0.23
Ast	0.998	0.985	0.998	0.979	0.07	0.20	0.07	0.24
Ska	0.998	0.964	0.998	0.982	0.06	0.28	0.06	0.18

directed flow: $d = \mathbf{U} \cdot \mathbf{U}_s / |\mathbf{U} \cdot \mathbf{U}_s|$, where \mathbf{U} and \mathbf{U}_s are the 2-dimensional horizontal velocity vectors from the unsteady and the steady model configurations respectively.

The direction parameter was then used to scale the velocity magnitude to separate ebb- and flood- directed flow velocities and time-averaged across a lunar cycle (denoted using angled braces). The results of the analysis (Fig. 7A and B; vector length does not represent magnitude), show that regions of ebb-directed net flow exist with velocities greater than 0.04 m s^{-1} (Fig. 7A). Zones of flood-directed net flow are observed in the lee of Grassy Island and along the channel margin (Fig. 7A). Increasing the river-discharge reduces the extent of these zones (Fig. 7B) and increases maximum tidal-averaged velocity (from

Table 2

Modelled and measured tidal amplitude and phase for the low (field-season 2, roman) and high (field-season 3, italics) discharge periods.

Gauge	Amplitude (m)											
	M ₂			K ₁			S ₂			O ₁		
Station	Obs	2D	3D	Obs	2D	3D	Obs	2D	3D	Obs	2D	3D
Ham	0.95	–	0.95	0.31	–	0.29	0.30	–	0.32	0.25	–	0.24
	<i>0.91</i>	–	<i>0.95</i>	<i>0.49</i>	–	<i>0.49</i>	<i>0.17</i>	–	<i>0.20</i>	<i>0.24</i>	–	<i>0.23</i>
Ast	1.01	0.97	1.02	0.30	0.29	0.30	0.30	0.30	0.32	0.23	0.23	0.24
	<i>0.96</i>	<i>0.91</i>	<i>0.97</i>	<i>0.47</i>	<i>0.46</i>	<i>0.46</i>	<i>0.16</i>	<i>0.16</i>	<i>0.18</i>	<i>0.22</i>	<i>0.21</i>	<i>0.21</i>
Ska	0.88	0.89	0.92	0.26	0.24	0.24	0.24	0.27	0.29	0.18	0.19	0.19
	<i>0.79</i>	<i>0.79</i>	<i>0.79</i>	<i>0.37</i>	<i>0.38</i>	<i>0.36</i>	<i>0.13</i>	<i>0.13</i>	<i>0.13</i>	<i>0.15</i>	<i>0.16</i>	<i>0.15</i>
Gauge	Phase (Degrees)											
	M ₂			K ₁			S ₂			O ₁		
Station	Obs	2D	3D	Obs	2D	3D	Obs	2D	3D	Obs	2D	3D
Ham	75	–	69	192	–	186	27	–	21	112	–	104
	<i>234</i>	–	<i>228</i>	<i>126</i>	–	<i>120</i>	<i>41</i>	–	<i>31</i>	<i>346</i>	–	<i>339</i>
Ast	87	89	81	198	201	192	41	44	34	115	120	108
	<i>247</i>	<i>250</i>	<i>243</i>	<i>133</i>	<i>137</i>	<i>130</i>	<i>61</i>	<i>63</i>	<i>49</i>	<i>357</i>	<i>1</i>	<i>356</i>
Ska	114	112	106	220	217	209	72	69	61	134	135	125
	<i>279</i>	<i>280</i>	<i>276</i>	<i>153</i>	<i>158</i>	<i>154</i>	<i>93</i>	<i>92</i>	<i>80</i>	<i>34</i>	<i>32</i>	<i>34</i>

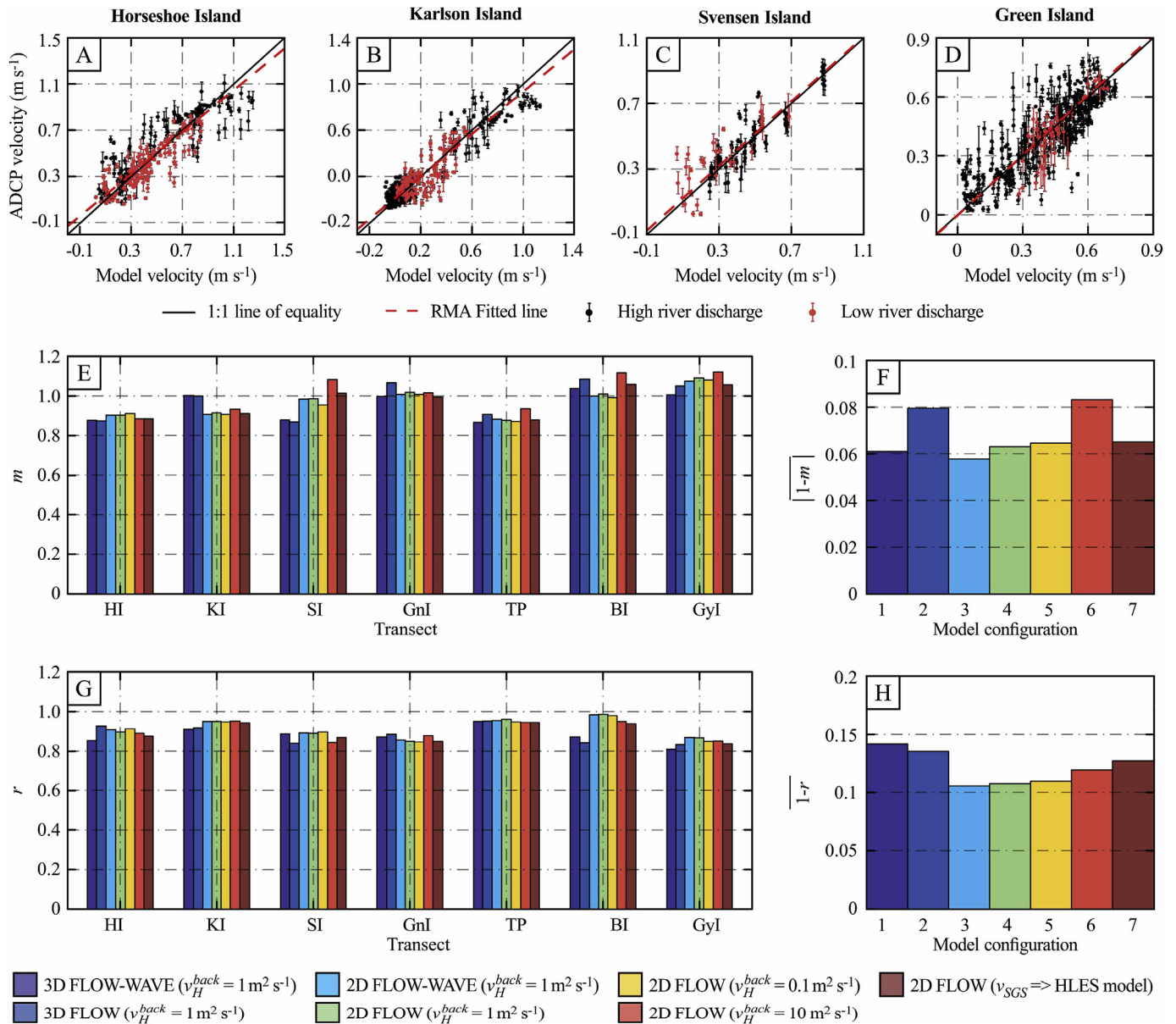


Fig. 5. (A–D) Comparison between measured ADCP and predicted model (2D FLOW $v_H^{back} = 1 \text{ m}^2 \text{ s}^{-1}$) velocities for four transects. Error bars represent one standard deviation from the within model-cell variability. (E & F) Gradient m obtained from RMA least-squares regression, (G & H) correlation coefficient.

0.18 m s^{-1} to 0.29 m s^{-1}).

These results indicate that there is potential for net sediment transport in this region. In principle, the sediment transport rate may be computed directly using a sediment transport scheme. However, application of these schemes requires careful site-specific calibration. For simplicity, the Engelund and Hansen (1967) transport scheme is used herein. A proxy for sediment transport is denoted with a lowercase s ($= U^5$). For the remainder of this paper, this is referred to as the sediment transport capacity. A similar approach has been adopted by others (e.g. Hoitink et al., 2003; Van Maren et al., 2004).

The sediment transport capacity was time-averaged over a lunar tide. The results are presented (Fig. 7C and D) as a fraction of local maximum tidal-averaged sediment transport capacity s_{\max} evaluated within region 1.2 (vectors denote the time-averaged direction) from the low discharge period.

During the high discharge period (Fig. 7D), sediment transport is nearly entirely ebb-directed with only small regions of flood-directed transport, in particular in the lee of Grassy Island. In the Prairie Channel, the ebb-directed transport is greater than the range of

contours presented, except for some areas close to the channel margin. During the low discharge period (Fig. 7C), sediment transport remains principally ebb-directed. In the lee of Grassy Island, and in some margins of the Prairie Channel, there are regions of flood-directed sediment transport with magnitudes greater than 5% of the local maximum. These regions of flood-directed transport have a larger spatial extent than the tidal-averaged velocity.

4.3.3. Tidal analysis of sediment transport in region 1.2

To understand the principal factors driving these flow and sediment transport patterns, the flood- and ebb- directed velocity magnitude were further analysed. The full signal is split into mean (U_0) and tidally-variable (U_T) velocities: $U = U_0 + U_T$. To do this: the tidally-variable part was computed using the `t_tide` Matlab toolbox (Pawlowicz et al., 2002) and the mean component obtained from the difference between the total and the tidally-variable velocity ($U_0 = U - U_T$). Using these reconstructed velocities, the mean (s_0) and tidally-variable (s_T) sediment transport capacities were computed.

The sediment transport capacities show that the tidally-variable part

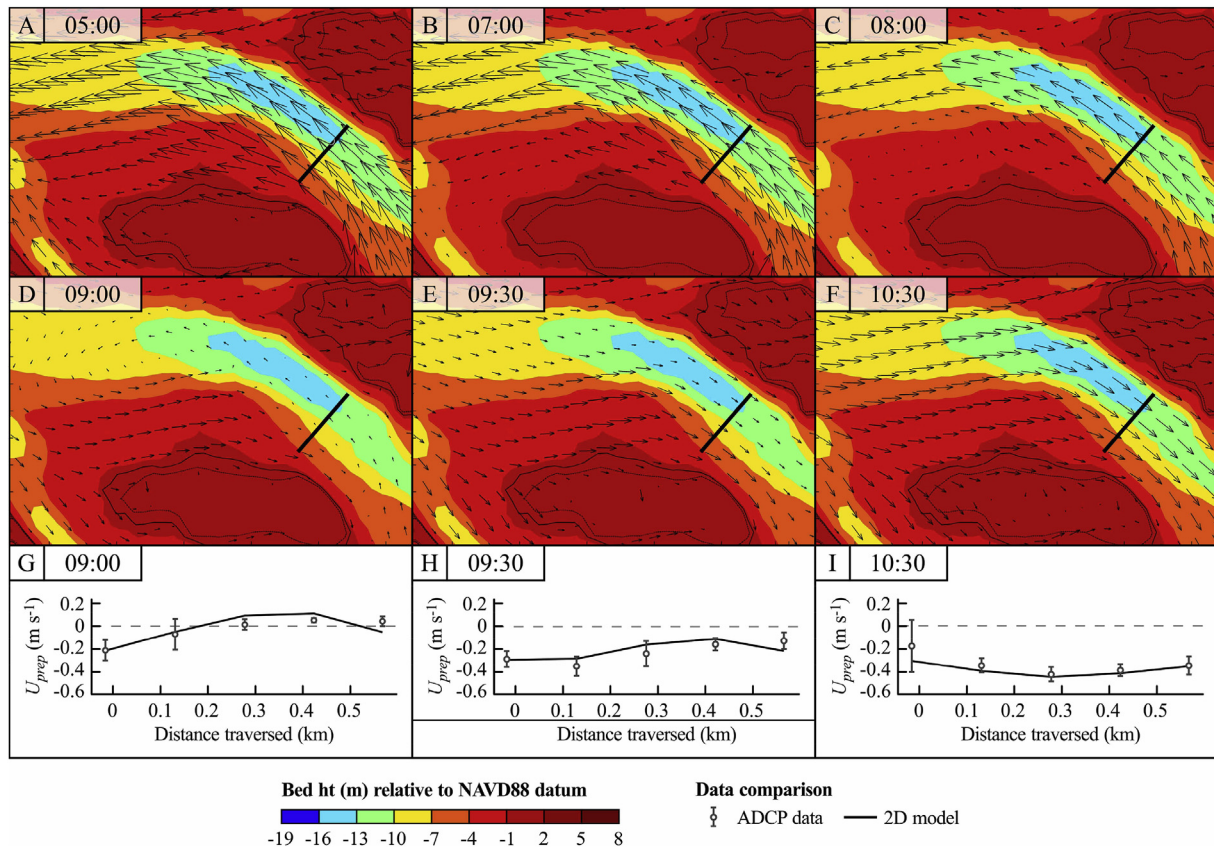


Fig. 6. Series of model and ADCP data showing the flow within a channel and across the Grassy Island transect (Fig. 1B) on the 18th June 2012 PST in region 1.1 (Fig. 1B).

(s_T) is significantly larger than the mean part (s_0) that is typically ebb-directed. In Fig. 7E and F, the tidally-variable sediment transport capacities are presented scaled by $s_{\max} = 0.0576 \text{ m}^5 \text{ s}^{-5}$. The tidally-variable part (s_T) is typically flood-directed with similar spatial distribution for both high and low discharge periods. The largest s_T values are aligned along the Prairie channel (denoted with dashed arrow) where water depths are largest in region 1.2.

To compare the total and reconstructed sediment transport capacities, we define the following ratios:

$$\begin{aligned} R_{sT} &= s_T / |s|, \\ R_{s0} &= s_0 / |s|. \end{aligned} \quad (2)$$

Negative values indicate flood-directed sediment transport.

The spatial distribution of R_{sT} are similar for both flow periods (Fig. 7G and H). The sediment transport is flood-directed and large compared to the total transport: i) along the north margin of the Prairie Channel and ii) in the layer between the Prairie Channel and the lee of Grassy Island.

Along the north margin of the Prairie Channel (i), the tidally-variable sediment transport appears to be the principal mechanism that results in flood-directed sediment transport.

Across the layer between the Prairie Channel and the lee of Grassy Island (ii), the total sediment transport capacity transitions from ebb-directed (within the Prairie Channel) to flood-directed (in the lee of Grassy Island) sediment transport. The large R_{sT} magnitude in this region is a result of relatively small total sediment transport due to competing tidal and mean flow. In the lee of the bar, the tidally-variable and mean flow sediment transport capacities are flood-directed (not shown) and so the total sediment transport capacity is flood-directed.

4.3.4. Tidal analysis of sediment transport in region 2

In Fig. 8A and B, the tidally-variable sediment transport ratio (R_{sT})

is presented for the upper to mid part of the estuary (region 2 – Fig. 1). Similar patterns are observed to those discussed previously. To compare these two ratios directly, the difference ($R_{sT} - R_{s0}$) is calculated and presented in Fig. 8C and D.

Analysis of the data showed that the dominant constituents are O_1 and M_2 for the low and high discharge periods respectively. The second largest constituent is K_1 for both discharge periods. Hoitink et al., (2003) showed that the linear interaction between these components (K_1 , M_2 and O_1) results in tidal asymmetry. They formulated analytical expressions to compute sediment transport (velocity raised to 3rd or 5th power). These formulations show that the long-term average sediment transport depends only on the amplitude and phase of these constituents.

To test these formulations in the CRE, we computed the sediment transport (5th power) using the obtained tidal components and Hoitink's formulation. The sediment transport patterns are very different to the sediment transport patterns computed using the time series data (not shown). In the CRE, the fluvial discharge is not negligible and thus the sediment transport capacity was also computed using the mean flow (M_0) combined with the same constituents as Hoitink et al., (2003) (K_1 , M_2 , O_1 , M_4). The resultant sediment transport patterns (not shown) are very similar to the results obtained using the original time series (Fig. 8E and F). We also tested these results with the Van Maren et al., (2004) formulation and similarly found that the sediment transport (this time a 3rd power) computed using time series data and reconstructed from components (M_0 , K_1 , M_2 , and O_1) were very similar.

4.3.5. Net sediment transport in region 2

Calculated sediment transport capacities are presented in Fig. 8E and F. The bed topography in this region is presented in Fig. 8G for reference. For the area presented, $s_{\max} = 0.5065$ and $2.6324 \text{ m}^5 \text{ s}^{-5}$ for the low and high discharge periods respectively. The maximum sediment transport capacities were obtained in the navigation channel in

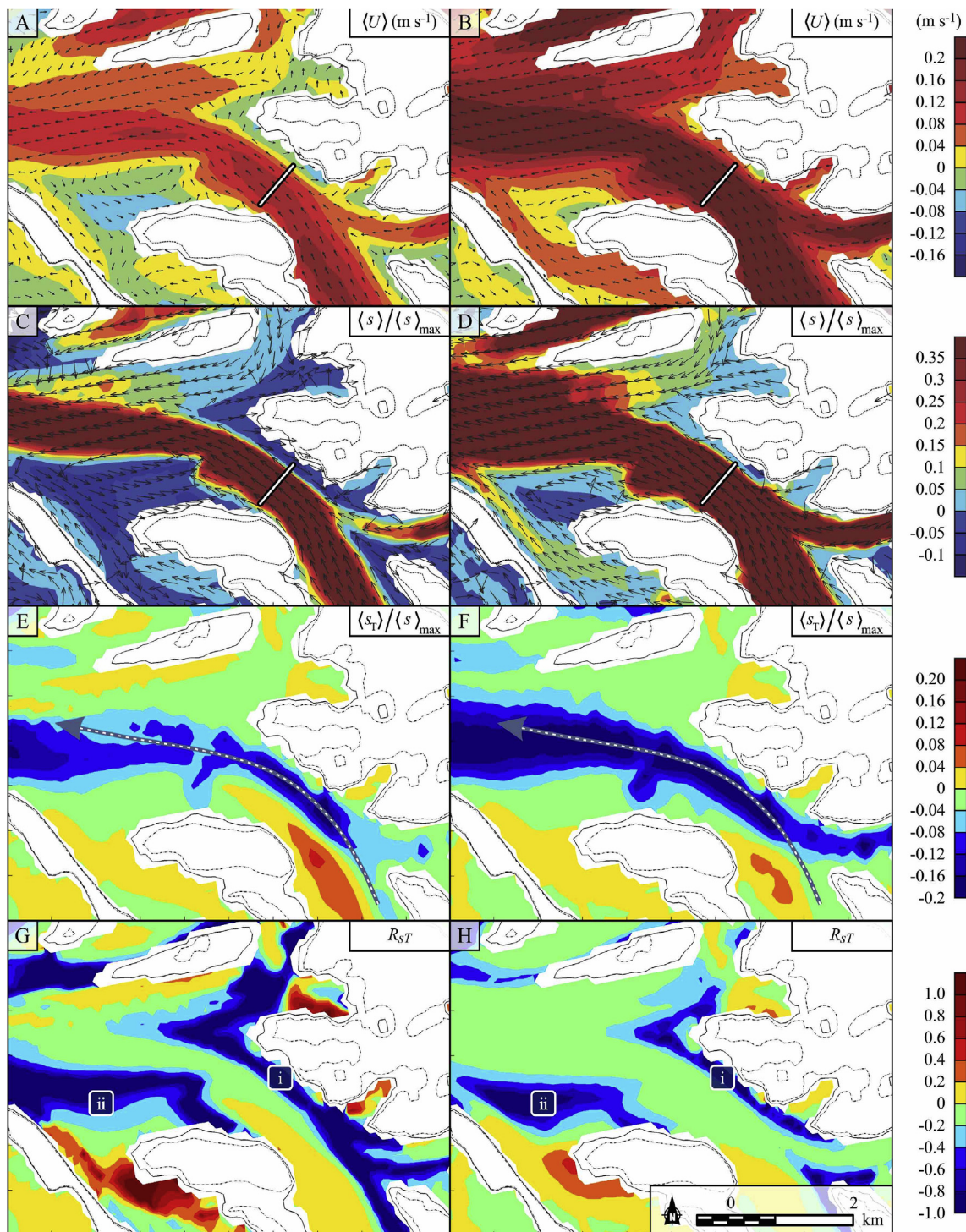


Fig. 7. Tidal averaged flow in region 1.2 defined in Fig. 1. (A & B) velocity and (C & D) total sediment transport, (E & F) tidally-variable sediment transport, (G & H) tidally-variable sediment transport ratio. ($\langle s \rangle_{\max} = 0.0576 \text{ m}^3 \text{ s}^{-5}$ local maximum during low river-discharge).

the upper part of the estuary.

In both the north and south embayments, most locations exhibit ebb-directed sediment transport around $0.05s_{\max}$ or less. In the remaining locations, small magnitude flood-directed and larger magnitude ebb-directed sediment transport is observed. The larger magnitude ebb-directed transport is concentrated in the deeper scoured areas. During low river flow periods (Fig. 8E), large regions of flood-directed sediment transport are observed in the lee of the bars and islands. Flood-directed sediment transport along the margins of the channels is also evident.

Finally, the ratio of sediment transport from the high and low river flow periods was computed (Fig. 8H). Sediment transport ratios greater than (less than) one indicate areas of increased (reduced) sediment mobility. The results show that this ratio is greater than one for most (92%) of the area presented. The ratio is spatially-variable with the largest values (~ 5) being located near the head of the estuary, and reducing along the axis of the estuary seaward (~ 2.5).

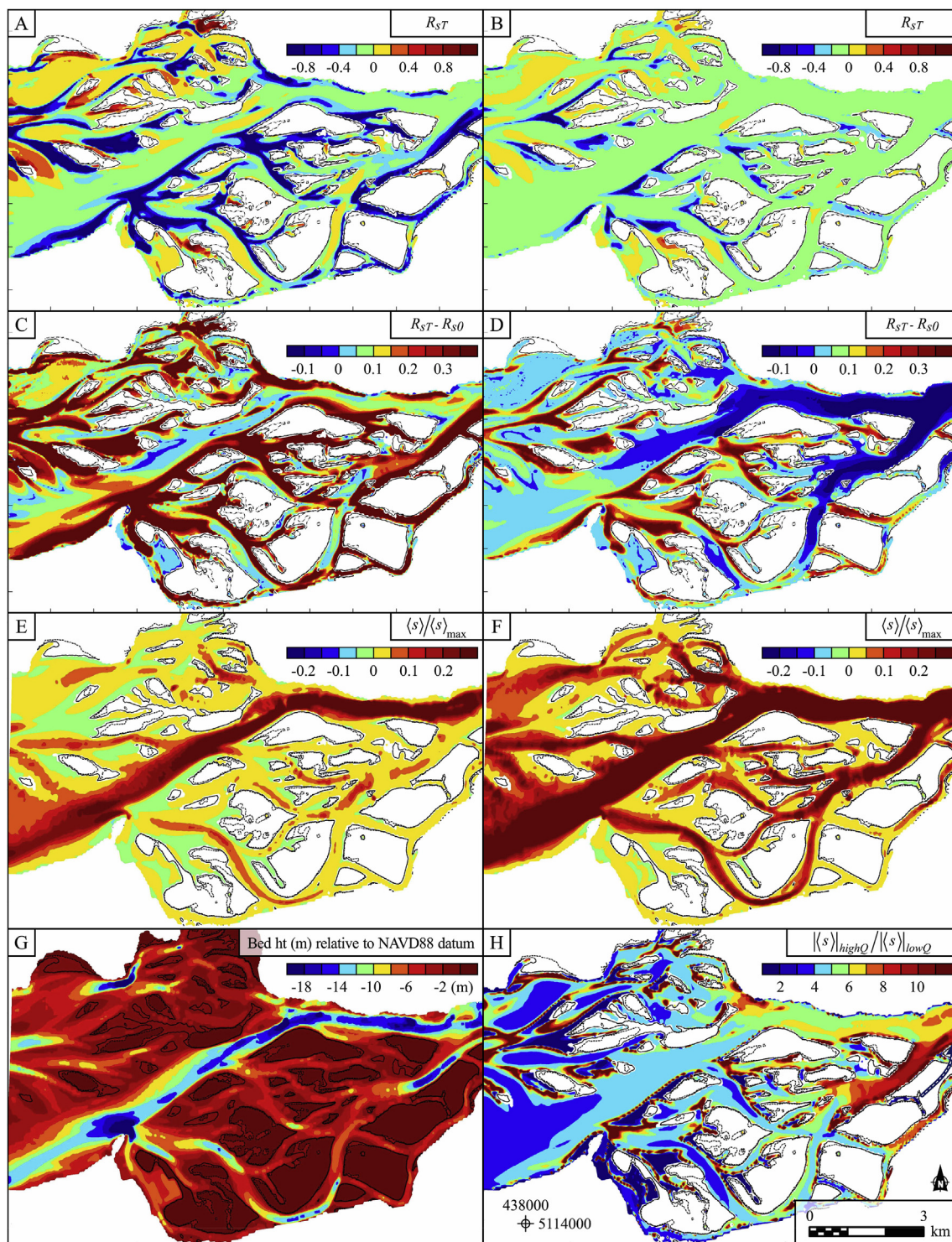


Fig. 8. Tidal average sediment transport in area 2. (A & B) tidally-variable sediment transport ratio, (C & D) difference between sediment transport ratios, (E & F) total sediment transport ($\langle s \rangle_{\max} = 0.5065 \text{ m}^5 \text{ s}^{-5}$ from the low discharge period), (G) bed level relative to NAVD88 and (H) ratio of the sediment transport from high and low discharge periods.

5. Discussion

The model calibration results presented herein highlight an inversion of the 3D model roughness sensitivity to water level variations due to a change in river-discharge. This was attributed to a change in the amount of work done during the two discharge periods. However, the 2D model did not exhibit this behaviour because the bed shear stress is related to the bulk momentum of the entire column in the 2D model

rather than being explicitly resolved.

From analysis of salinity and water level data (not shown), we found that the 3D sea-boundary model performed to a similar level of accuracy compared to other models (Elias et al., 2012; Kärrnä et al., 2015; Kärrnä and Baptista 2016) of the CRE. Our results also demonstrated that the 2D estuary-boundary configuration performed better than the 3D sea-boundary configuration in terms of calibration and validation statistics. This was based on analysis of observed water-level and flow

data. This may be due to modelling the more complex system and including coastal and mouth processes, and demonstrates the difficulties associated with modelling the CRE. One particular difficulty is related to the specification of the seaward boundary. In the approach adopted herein and by Elias et al., (2012), the sea boundary height is specified using stationary mean water level. The associated error in the estimation of this height is 0.226 m (VDatum: http://vdatum.noaa.gov/docs/est_uncertainties.html). In the 2D estuary-boundary simulations, the water levels are specified using measured data, and the model boundary conditions are therefore more constrained. Our results suggest that, although stratification and vertical process representation are important, similar if not better model predictions can be obtained using a 2D model constrained within the estuary.

The differences in performance may also be due to running the 3D model at lower resolution. As a test, we ran additional simulations at the lower resolution for the 2D case. Reducing the resolution results in slightly poorer calibration and validation statistics although the differences are small (not shown). We therefore expect that the 3D model calibration and validation statistics to be only marginally better if run at higher resolution.

The temporal and spatial flow data show that larger peak velocities are associated with the deeper areas of the channels, likely due to topographic forcing. The associated increase in momentum results in a greater lag in response to water-level change compared to comparatively slower flow in the shallower and more sheltered areas. This was demonstrated in an area of the south embayment (region 1.1). This results in net ebb-directed flow in the deeper channels and flood-directed flow into the shallower and more sheltered areas (region 1.2). Similar spatial distributions were obtained for net-sediment transport capacities (region 1.2). Increasing the river-discharge results in a reduction in the spatial extent of the flood-directed flow and sediment transport.

Although the sediment transport is principally ebb-directed, there are areas of net flood-directed sediment transport, which has two implications for sedimentological processes. Firstly, in the lee of islands and bars, flood-directed flow provides a source of sediment that could subsequently be deposited, and a mechanism to extend bar/island topography downstream. Secondly, flood-directed flow may produce flood-directed bedforms. If these bedforms are preserved in bar/island stratigraphy, they could provide a signature that may help interpret the origin and evolution of depositional forms in the TIFZ. However, their preservation potential is also linked to aggradation rate, and thus in regions of large amplitude gross-, but low amplitude net-, sediment transport, the bedforms produced are likely to be more symmetrical. Such a situation is observed at the margins of Prairie Channel and in the lee of Grassy Island during high river-discharge. At the same time, large ebb-directed sediment transport will likely produce ebb-directed bedforms. This yields the potential for spatial variations in the degree of bedform symmetry and therefore the potential difficulty in associating certain bedform assemblages with a downstream position in the fluvial-tidal zone (Venditti et al., 2012).

To better understand the principal driving mechanisms for flood- and ebb-directed sediment transport, the tidal signal was split into tidally-variable and mean parts. To compare the influence of each, two parameters were used: R_{sT} and R_{s0} (Equation (2)). The tidally-variable part (R_{sT}) is principally flood-directed, large compared to R_{s0} and spatially variable. In particular, large negative (flood-directed) R_{sT} is observed in two locations: i) along the margin of the channels where tidal variability is the main mechanism for flood-directed sediment transport (R_{s0} is typically ebb-directed); and ii) in areas where the principal net sediment direction changes (in the lee of bar topography), this highlighting areas where the mean part is also negative. The converse is true for the mean part R_{s0} (not shown). In region 2, R_{s0} is fairly uniform and generally small compared to $< s >$ during the low discharge period. However, during the high discharge period, R_{s0} exhibits greater variability increasing in magnitude towards the head of the estuary.

Furthermore, direct comparison between these parameters shows that the tidally-variable ratio is dominant - in particular during the low discharge period and along the channel margins.

Although R_{s0} is generally small compared to R_{sT} , the mean component is crucial in determining the resultant sediment transport patterns. This was observed in tests where the resolved constituents without the mean flow (M_0) were used to reconstruct the sediment-transport (not shown here). Furthermore, given K_1 , M_2 and O_1 are the dominant constituents in the CRE and have previously been linked to tidal asymmetry (Hoitink et al., 2003), we also tested the following combination: K_1 , M_2 , O_1 , M_4 plus M_0 . The spatial sediment transport patterns were computed by reconstructing the flow from the given constituents. The reconstructed sediment transport patterns are visually similar to the total sediment transport, and demonstrates the importance of the mean flow in determining the sediment transport patterns in the CRE. This was also demonstrated by Chu et al., (2015) who found that the mean flow can modify the temporal, and therefore long-term, transport patterns.

The impact of river-discharge on the sediment transport patterns in the CRE is dominated by three factors. Firstly, due to topographic forcing, river-discharge is preferentially routed into deeper parts of the channels. This results in ebb-dominance in the deeper channels. Flood-dominance is still possible, and was demonstrated in the shallow and sheltered areas, but increasing river-discharge limits the spatial extent of the flood-directed areas. Secondly, although the magnitude of the mean flow can be small, its effect on determining the sediment transport patterns is significant. Finally, the effect of increasing the discharge leads to increased sediment transport.

6. Conclusions

Delft3D was applied to simulate the hydrodynamics of the Columbia River Estuary, with testing of two modelling configurations and parameter sensitivity. The models were calibrated using water level data and validated using salinity and flow data. Overall, the model performs very well at reproducing the measured water levels and flow patterns. Our results show that a 2D estuary-boundary model can be used to provide a good representation of planimetric hydrodynamics within a highly energetic and stratified estuary.

A detailed description of the flow within an area of the south embayment illustrates the complex interactions between hydrodynamic forcing of the tidal flood with the momentum of the ebb-directed flow during the turn of the tide. Flow reversals occur first in areas sheltered by bar/island topography and along the margins of the channels, thereby producing planimetric bidirectional flows and recirculation zones.

These interactions result in net sediment transport that is principally ebb-directed. However, within the channels and sheltered areas, flood-directed flows and sediment transport are evident. The results indicate that this is due to topographic forcing that focuses most of the river-discharge into deeper parts of the channels. Increasing the river-discharge results in a reduction of the extent of the flood-directed flow and sediment transport. Further analysis of the results showed that: i) flood-directed sediment transport along the channel margins is due to tidal variability, and ii) flood-directed sediment transport in the lee of bars is due to both tidal variability and mean flow.

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